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by

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EARLY WINTER IN THE NORTHERN STRATOSPHERE

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SUMMARY

An account is given of the evolution and structure of the circulation in the stratosphere and upper troposphere of the northern hemisphere in the early winter of 1982/83. An early winter warming is described and it is shown to have some features in common with Canadian warmings. A tentative connection is made between the intensification in the troposphere of a trough/ridge pattern associated with the East Asian Low and the establishment of the Aleutian High as a persistent feature extending through the whole depth of the stratosphere. Processes leading to the formation of the so-called surf-zone are considered, and aspects of zonal-mean diagnostics are related to developments on synoptic maps.

1. INTRODUCTION

The winter stratosphere in the northern hemisphere is dominated by a huge cyclonic vortex and strong, planetary-scale disturbances to it. A common feature on synoptic maps is the anticyclone referred to as the Aleutian High. Its appearance as a persistent vortex extending right through the stratosphere is often presaged in early winter by a distinct type of warming during which the flow is disrupted in the lower and middle stratosphere but not in the upper stratosphere (unlike minor warmings which mostly affect the middle and upper stratosphere). These warmings vary in intensity from one year to the next but have a number of features in common. Strong warmings of this type are called Canadian warmings by

Labitzke (1977). During such warmings, warm air moves polewards and eastwards across Canada. The cyclone is pushed off the pole and is highly distorted, but it stays strong whereas in a major warming it breaks down.

There are only a few studies of early winter in the stratosphere. Labitzke et al. (1979-1986) summarized, for several northern winters, the evolution of the stratosphere during this period using data from radiosondes and rockets, and Clough et al. (1985) used data from satellites to discuss the particulary intense Canadian warming of 1981/82. Our intention is to extend the latter work. We study the early winter of 1982/83 with the following aims:

1. To describe the evolution and vertical structure of the circulation in the stratosphere during the early winter of 1982/83, noting features of a warming that are in common with the Canadian warming studied by Clough et al.

2. To identify features of the tropospheric flow which might exert a significant influence on the flow in the stratosphere.

3. To determine how and to suggest why a structure that has been called a surf-zone comes into being.

4. To relate the evolution of zonal-mean winds in the stratosphere to changes on synoptic maps, thereby judging the extent to which the theory of wave, mean-flow interaction is appropriate.

We shall use maps of Ertel's potential vorticity, Q, a quantity defined thus:

$$Q = \rho^{-1} \, (\underline{\nabla} \times \underline{u} + 2\underline{\Omega}) \cdot \underline{\nabla}\theta \qquad , \qquad (1)$$

using standard notation as detailed in the Appendix. Q has two properties that make it of fundamental value in a dynamical analysis. (1) For adiabatic inviscid motion, it is conserved following an air parcel, as is potential temperature. Contours of Q on isentropic maps are then material lines which can be used to trace the movement of air. (2) Under balanced conditions (e.g. thermal wind balance) and with suitable boundary conditions, all other fields can, in principle, be derived from the three-dimensional distribution of Q (Hoskins et al., 1985). In an observational study, Q is derived from other fields not vice versa, but used together these properties are a conceptual aid in analysing the dynamics of large-scale flows. From (1), a local anomaly in Q can arise only by advection of air with that Q from somewhere else on an isentropic surface. From (2), we may think of high or low Q as being a source of cyclonic or anticyclonic circulation. Hence, the Aleutian High builds and is replenished against dissipation by advection of low Q from low to middle latitudes; and the main cyclone shrinks partly by being drained of its high Q by the Aleutian High, as McIntyre and Palmer pointed out (1983, 1984, 1985).

Use of property (1) is limited by radiative heating or cooling, and by the development of features too small to be observed. Provided that the flow is adequately resolved, a contour of Q can be taken as a material line

for about a week or so in the mid stratosphere, where the time for radiative damping is a few weeks for deep perturbations of the kind we shall study. Small scales can arise when large-scale velocity patterns stretch out tongues of Q, as in Fig. 1 (see caption). The importance of small scales has not been quantified for the stratosphere, but McIntyre and Palmer have proposed that irreversible mixing of Q is significant during what they call planetary-wave breaking (not the overturning of fluid elements in the vertical, but the irreversible buckling of material lines on an isentropic surface).

The plan of this paper is as follows. In section 2, we describe the data used and how we processed them to derive fields of Q. Section 3 contains our synoptic description of early winter, 1982/83. We consider in section 4 systematic changes in the structure of the circulation using as a diagnostic the areas contained within contours of Q for various values of Q. In section 5, we discuss how the evolution of zonally averaged quantities is related to advection of Q, and our conclusions are given section 6.

2. DATA AND METHOD OF ANALYSIS

The data used here were obtained mainly from stratospheric sounding units on board the satellite NOAA-6 of the Tiros-N series. Details of data retrieval and a discussion of the errors involved are given by Clough et al. (1985) and references therein. Briefly, Q is calculated as follows:

1. Radiances are inverted to give thicknesses by statistical regression.

2. The thicknesses are added to an analysis of geopotential height at 100 mb provided by the National Meteorological Center (Washington), giving geopotential heights at 20, 10, 5, 2, and 1 mb. An NMC analysis at 50 mb is included to supplement the vertical resolution in the lower stratosphere.

3. Fields are smoothed by Fourier analysis with truncation at wavenumber 12 in the zonal and meridional (pole to pole) directions.

4. Geostrophic winds are calculated (gradient winds give noisier fields of Q).

5. Q is computed using an approximation valid for small Rossby number and large Richardson number:

$$Q \approx - g (\zeta_p + f) \partial\theta/\partial p \qquad (2)$$

6. Fields of Q and winds are interpolated on to isentropic surfaces.

Because of the smoothing inherent in the measurements and method of analysis, the isentropic maps of Q thus obtained give only a blurred view of the distribution of Q in the stratosphere. Smoothing in the horizontal

limits resolution to scales greater than about 30° in longitude and about 15° in latitude. Since the stratosphere is sampled in slabs, these estimates apply only to features deeper than about 10 to 15 km.

3. SYNOPTIC DESCRIPTION OF EARLY WINTER, 1982/83

Radiative cooling in the polar night creates a strong circumpolar westerly vortex in the stratosphere during early winter. In the winter of 1982/83, the first hint of an Aleutian High in the middle stratosphere was in mid October, but it was a weak, transitory feature. By the end of December, however, it was firmly established through the whole depth of the stratosphere. How did this happen?

Fig. 1 is a map of Q and geostrophic wind for 5 December on the 850K isentropic surface, lying near 10 mb in the mid stratosphere. The Aleutian High is centred on a bubble of low Q that has been injected from low into middle latitudes (a sequence of such maps confirms that movement of air rather than non-conservative processes is involved in creating the bubble). The values of Q in the bubble imply that air has made a long excursion to high latitudes from south of about 30°N. The Aleutian High strongly and irreversibly deforms the material lines in the westerly vortex, a characteristic of planetary-wave breaking according to McIntyre and Palmer (1983, 1984, 1985). That is, some of the Q contours are not simply undulating back and forth as they would in the presence of linear waves, but are stretched out into long tongues (such as the one shaded in the figure). Although there is a strong anticyclonic circulation around the Aleutian High, the tip of the tongue of high Q is in a region where winds

are weak, so if high Q is entrained into the Aleutian High and mixed therein, this happens only slowly. We shall record the fate of this tongue in due course.

An analytical solution that has been used (e.g. by McIntyre and Palmer, 1984) as an analogue of such wave breaking events is that found by Stewartson (1978) and by Warn and Warn (1978) for Rossby critical layers. This solution predicts that the wave will break at a zero-wind line (if the zonal phase speed is zero, as it nearly is in the present case). We find that there is no zero-wind line north of 20°N (the southern limit of our analysis of zonal wind) at 10 mb on 5 December; indeed, the anticyclone is centred not where winds are weak but close to the maximum of the zonal-mean jet. Clough et al. (1985) and O'Neill and Pope (1988) recorded other examples of wave breaking in the stratosphere well away from a zero-wind line. In such cases, the fall in density with height, leading to an increase with height of disturbance geopotential, seems to be a controlling factor.

Since it is well known that large-scale disturbances in the stratosphere are forced in the troposphere, we look there for the origin of the Aleutian High. Fig. 2 shows the field of geopotential height at 100 mb on 5 December (100 mb lies in the lower stratosphere, but the flow resembles closely that of the upper troposphere except for small-scale disturbances that do not propagate vertically). From a later sequence of such maps, we infer that the Aleutian High grows in unison with the trough shown in Fig. 2 along 135°E and a downstream ridge which develops over the Pacific. (The strongest winds in the hemisphere are on the southern flank

of the trough.) Such an association is reasonable in the light of the idealised experiments reported by O'Neill and Pope (1988). They used a primitive equation model of the stratosphere and mesosphere whose lower boundary near the tropopause is prescribed. Applying a geopotential "low" to an otherwise symmetric field at this boundary, they found that an anticyclone was generated in the mid stratosphere about 40° longitude downstream from the perturbation. The corresponding features in Figs. 1 and 2 are arranged similarly. The trough in the troposphere over East Asia and downstream ridge appear on time-mean charts in winter, as does the stratospheric Aleutian High.

Fig. 3 is the map of Q and winds on 13 December. The bubble of low Q persists near 120°W, but values of Q in it gradually increase, and the attendant anticyclone weakens (the contour labelled "2", which surrounds the bubble in Fig. 1, is no longer there in Fig. 3). A systematic increase in a local minimum in Q can arise only by a non-conservative process (among which we include the generation of unresolved motions); advection would merely move that minimum somewhere else. In section 4, we consider some possible candidates. The Aleutian High does not decay completely, however, because it is reinforced by advection of low Q around the edge of the displaced main vortex (notice the tongue of low Q in the top right quadrant of Fig. 3). As the parcel moves polewards, its anticyclonic vorticity increases. This may be deduced from Eq. (2) and the fact that a deep structure does not alter its static stability much, so that an increase in planetary vorticity is balanced by an increase in anticyclonic relative vorticity. The anticyclonic circulation is at a maximum when the parcel reaches the location of the Aleutian High.

The fresh injection of low Q marks the onset of a warming that shares a number of features with the kind of warming referred to by Labitzke (1977) as a Canadian warming. Invoking the invertibility principle, we may regard the high Q in the large cyclone as inducing an eastward component on the winds at the tip of the tongue of low Q. The tongue is consequently drawn eastwards over Canada (west coast near 125°W), as shown in Fig. 4 for 18 December.

The event of early December 1982 reproduced two of the salient aspects, noted by Clough et al., of the Canadian warming of 1981, though the latter disturbance was much more intense. (1) There is little tilt with height of the disturbance at the peak of the warming. A sequence of maps at different levels indicates that initial westward tilt is reduced as low Q extends eastwards in the stratosphere, while the trough in the troposphere over E. Asia does not move. The barotropic structure is consistent with the modest increases in temperature (about 20K). (2) The westerly circulation is disrupted only in the lower and middle stratosphere; there is no closed anticyclonic circulation at mid latitudes in the upper stratosphere. A characteristic of Canadian warmings according to Labitzke (1977) is that the amplitude of the thermal disturbance decays with height in the middle stratosphere.

In major warmings, on the other hand, typical temperature increases are about 80 K, and there is at least one strong anticyclone in the upper stratosphere (usually the Aleutian High). Presumably, this difference between warmings stems, in part, from differing tropospheric conditions.

The main disturbance in the troposphere during the warming of early December 1982 was the trough over E. Asia (the flow was similar to that shown in Fig. 2), but it was weaker and less persistent than the troughs and ridges that build in the troposphere during major stratospheric warmings.

The tongue of high Q persists at low latitudes through most of December (appearing on a monthly-mean map), though values of Q in it decrease (compare Figs. 1 and 4). During the last week of December, however, the Aleutian High intensifies and Q is rearranged, though our view of what happens is fuzzy. Fig. 5 is the map of Q and winds for 30 December. The pattern of adjacent tongues is destroyed. Apparently, the growing anticyclone draws high Q from the tongue at low latitudes back to high latitudes, thereby isolating a patch of low Q on which the anticyclone is centred. Notice that the winds blow across the contour labelled "2" near 30°N, 150°W (the patch of low Q is not quite isolated from similar values of Q at low latitudes). In principle, there is no reason they should not do so; but in this case, it is more likely that the position of this contour is uncertain because of vertical structure which is not fully resolved. The stratosphere is more baroclinic than it was hitherto, and the Aleutian High is now present at mid latitudes up to the stratopause.

The troposphere was more disturbed at the end of December than it was during the warming in early December. Fig. 6 shows the geopotential height at 100 mb on 30 December. The East Asian Low is deeper, and the downstream ridge and overlying Aleutian High are more intense.

By the end of December, the stratosphere contains the kind of features that McIntyre and Palmer (1984) call a main vortex and a "surf-zone". The main vortex is defined by a strong gradient of Q, its edge by a sharp transition to a region where the gradients are much weaker - the surf-zone. Accordingly, in Fig. 5, the shading roughly defines the edge of the vortex, and the Aleutian High resides in the so-called surf-zone. Butchart and Remsberg (1986) use a numerical model to demonstrate that such a structure does not arise in a stratosphere which undergoes a seasonal cycle without planetary-scale disturbances. Non-conservative processes associated with such disturbances are therefore implicated. We now discuss what these might be.

4. SYSTEMATIC CHANGES AND NON-CONSERVATIVE PROCESSES

As the Aleutian High is established in early winter, there are systematic changes in the structure of the circulation. These changes are summarized by computing, on an isentropic surface, the total area where Q is greater than a specified value for several values of Q. This procedure was recommended by McIntyre and Palmer (1984), and was adopted by Clough et al. (1985) and by Butchart and Remsberg (1986). Assuming that the large-scale flow is approximately non-divergent, these areas would not change in the absence of non-conservative processes (in a non-divergent flow, material lines enclose a constant area). They do change, however, so such processes must be at work. We can think of three potential candidates: (1) radiation; (2) loss of resolution, which may be irreversible; (3) turbulence due to breaking gravity waves. To discuss them further, we first examine the seasonal changes in the areas

circumscribed by Q contours. We show results for a period longer than that of our earlier synoptic discussion, first to distinguish long-term trends from short-term variability, and second to discriminate between the processes where possible.

Fig. 7 shows how the areas change from July 1982 to June 1983. The curves are labelled with the relevant value of Q in the units given in the figure legend, and we shall denote the areas by A(Q) with Q in these units. The areas increase steadily from summer to early winter as the westerly vortex builds by radiative cooling of air in the stratosphere. From about mid November, however, the areas shrink within Q contours farthest from the centre of the vortex (those labelled "2" and "3" on the figure), while areas for Q contours nearer the centre keep rising. Such a difference between inner and outer contours would not arise in an axially symmetric, 'reference' stratosphere whose seasonal cycle was governed only by the solar elevation (Butchart and Remsberg, 1986, Fig. 6). As westerlies strengthen in the stratosphere, planetary-scale disturbances can propagate upwards from the troposphere (Charney and Drazin, 1961). They tend to be refracted towards low latitudes (Matsuno, 1970), and this is where the first sustained departures from the reference cycle arise.

A(3) decreases rapidly from mid November till the end of December, whereas A(2) increases in late December and continues to do so till the end of February (apart from short-term fluctuations). The average separation between Q contours "2" and "3" on an isentropic map increases systematically with time. At the end of December, the signature of the so-called main vortex and surf-zone is already apparent in the gradient,

dA(Q)/dQ: the gradient is large for high Q and small for low Q; and there is a sharp change in gradient in between. Accordingly, Q contour "3" might be deemed to lie on average in the main vortex, and Q contour "2" in the "surf-zone".

First, we ask why A(3) starts to decrease before mid winter (in the 'reference cycle' the decrease begins slightly after the winter solstice). A possible cause is radiation arising when disturbances remove the atmosphere from radiative equilibrium. Butchart (1987), and O'Neill and Pope (paper in preparation) demonstrate that such radiation can decrease areas enclosed by Q contours associated with the main vortex, but by less than is observed (O'Neill and Pope note, however, that the calculations are prone to large errors when based on observational data).

Another possibility is that Q is lost through inadequate resolution. In Fig. 1 for 5 December, Q contour "3" lies partly in a long tongue (shaded). The wind arrows imply that the tongue will become longer, so it may also become thinner if the circulation in the tongue is approximately non-divergent and thus area conserving. By 18 December, Fig. 4 shows that values along the tongue are generally less than 3. We can estimate the area, A(3), that could be hidden from view by inadequate horizontal resolution (we assume vertical resolution is adequate). The tongue defined by Q contour "3" lies roughly along latitude 40°N; it must be less than about 15° wide in latitude or it would be seen; and, from Fig. 4, we shall assume that it stretches 90° around the latitude circle. Then we find that the hidden area can amount to about 10^{13} m^2. This can account for about half the decrease in A(3) between the beginning and end of December (see

Fig. 7). The rest might be accounted for by a combination of mixing into the Aleutian High and radiation. Values of Q down the middle of the tongue are very close to 3, and only a small error is needed to change A(3) this much. Q contour "4" is also part of a tongue in Fig. 1, and a similar calculation could explain the drop in A(4) in early December. Contours of higher value than 4 are not drawn into tongues in early winter, and the areas they contain change more like those in the reference cycle.

Finally, there could be small-scale mechanical effects acting to reduce Q. Drag produced by dissipating gravity waves is usually assumed to be large only in the mesosphere (Lindzen, 1981), but as far as we know there are no reliable direct measurements of the drag in the middle stratosphere.

Next we ask why, apart from short-term fluctuations, A(2) increases systematically from the end of December till the end of February (when in the reference cycle it decreases). McIntyre and Palmer (1983, 1984) proposed that planetary waves breaking in the stratosphere would mix potential vorticity around the outside of the westerly vortex. They suggested that the Aleutian High may twist up Q contours in a way analogous to what happens in a certain analytical theory (developed by Stewartson, 1978, and by Warn and Warn, 1978) when Rossby waves impinge on a zero-wind line. According to this hypothesis, weak gradients of Q would develop in the Aleutian High because Q would appear mixed in a blurred view (the wind field need not be chaotic for this mixing to occur). If high values of Q

are twisted up with low values of Q, then A(Q) decreases for high Q and increases for low Q. The evolution of A(2) and A(3) in the second half of December is consistent with McIntyre and Palmer's hypothesis.

There is indirect evidence in Fig. 7, however, that radiation plays an important role in weakening the average gradient of Q around the westerly vortex in the stratosphere. Notice that in mid February, when a strong warming begins, A(3) starts to increase markedly (apart from short-term fluctuations). If this increase were due to mixing alone, we would expect that the areas for higher values of Q would start to decrease more sharply than before, which they do not do (again, apart from short-term fluctuations). Calculations by O'Neill and Pope (paper in preparation) confirm that radiation plays a significant role in effecting systematic changes like the ones noted here, in particular in establishing the weak gradient of Q that has been taken (McIntyre and Palmer, 1983, 1984; Butchart and Remsberg, 1986) as the hallmark of a 'surf zone'. But they also find that, as McIntyre and Palmer hypothesized, loss of resolution has a significant effect on the measured distribution of Q. We reached the same conclusion on the basis of the rough calculation given above.

5. ZONAL-MEAN DIAGNOSTICS

Diagnostic studies of stratospheric sudden warmings have often been based on the theory of wave, mean flow interaction. The theory, as used in applications to date, relies on two principles:

(1) The divergence of the Eliassen-Palm flux gives the effective force per unit mass acting on the zonal-mean wind. This result is exact for any disturbance amplitude.

(2) Waves can be treated as propagating quasi-linearly on the zonal-mean state, which is assumed to evolve much more slowly than do the waves.

Principle (2) is not satisfied for the early winter warmings of 1982/83. Disturbance amplitudes are too large for linear theory to be an acceptable approximation, and (related to this) the time-scale for changes in the zonal-mean winds is the same as that for the disturbances. The correlation that we and others have found between the divergence of the Eliassen-Palm flux and the acceleration of the zonal wind gives a good check on the quality of the data, but does not, in any rigorous sense, allow stratospheric warmings to be explained causally in terms of waves 'forcing' changes in a mean flow.

In this section, we relate the evolution of zonal-mean diagnostics to the synoptic evolution described above, with the aim of clarifying aspects of strong warmings noted in earlier studies.

Fig. 8 shows the evolution of zonal-mean winds in the upper stratosphere at 1 mb during December 1982 and early January 1983. At the start of the period, there is a broad, strong zonal-mean jet. Subsequently, zonal-mean winds accelerate at high latitudes and decelerate at low latitudes, resulting in a poleward movement of the zonal-mean jet

and a narrowing of its latitudinal extent. At the end of December, as the Aleutian High is established in the upper stratosphere, zonal-mean winds decelerate rapidly at mid and high latitudes.

The polar movement of the zonal-mean jet has been regarded as a pre-condition for a strong warming (e.g. Kanzawa, 1982). It is supposed that waves propagate quasi-linearly on the zonal-mean jet and are focused by the high-latitude, narrow jet to give a strong polar warming. Indeed, we find that there is strong convergence of the Eliassen-Palm flux at high latitudes in the stratosphere at the end of December.

We do not, however, interpret our observations in terms of linear wave dynamics. In an idealised simulation of a stratospheric warming with a numerical model, O'Neill and Pope (1988) found that the simulated zonal-mean jet in the stratosphere moved poleward as the warming progressed, as did the zone where Eliassen-Palm fluxes converged. Adapting arguments given by Stewartson (1978), Warn and Warn (1978), and Killworth and McIntyre (1985), they contended that this behaviour was due to nonlinear, isentropic advection of potential vorticity in a widening 'buckling zone' - a region where material lines (Q contours) are deformed irreversibly. During early winter 1982/83, the buckling zone encompasses the Aleutian High, this being where Q contours are most strongly deformed. As the Aleutian High develops, the buckling zone widens and the region of Eliassen-Palm flux convergence moves to high latitudes. In line with O'Neill and Pope's findings, zonal winds strengthen at high latitudes as a localised jet stream forms between the Aleutian High and the westerly

vortex (Fig. 5). When the buckling zone is as wide as it is shown to be in Fig. 5, there is no well-defined region in the winter stratosphere where linear wave theory is applicable.

6. CONCLUSIONS

In an order matching the list of aims given in section 1, our conclusions are these:

(1) The stratospheric warming of early December 1982 belongs to a distinct class of early winter warmings. It strongly disrupted the circumpolar westerly flow in the lower and middle stratosphere, but the upper stratosphere went largely unaffected. The warming had a barotropic structure, consistent with the modest changes in temperature that occurred. These features are in common with the Canadian warming of early winter 1981/82 studied by Clough et al. (1985).

(2) As far as we could judge diagnostically, the warming was the stratospheric response to the establishment in the troposphere of the climatological East Asian Low and its attendant downstream ridge. When these features intensified and became persistent, the Aleutian High also strengthened and persisted. The closed anticyclonic circulation associated with it then extended through almost the whole depth of the stratosphere.

(3) We proposed that systematic changes in the average distribution of potential vorticity in the stratosphere resulted (a) from loss of resolution as potential vorticity was drawn out into thin tongues by the Aleutian High, in the manner first noted by McIntyre and Palmer (1983), and (b) from the non-conservative effects of radiation acting on a circulation persistently disrupted by planetary-scale disturbances. These processes, acting in concert, led to the appearance in the stratosphere of the northern hemisphere of a so-called surf-zone by the end of the year, 1982.

(4) We related the poleward movement of the zonal-mean jet as the Aleutian High was established, and the attendant poleward migration of the zone of Eliassen-Palm flux convergence, to nonlinear advection of potential vorticity in a widening 'buckling zone' - where Q contours are deformed irreversibly - rather than to a wave, mean-flow interaction and linear wave propagation.

APPENDIX

f	Coriolis parameter
g	acceleration due to gravity
p	pressure
Q	Ertel's potential vorticity
$\underline{u}$	three-dimensional velocity vector
ζ_p	relative vorticity on an isobaric surface
$\underline{\nabla}$	three-dimensional gradient operator
θ	potential temperature
$\underline{\Omega}$	Earth's rotation vector
ρ	density

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FIGURE LEGENDS

Figure 1. Polar-stereographic map of Ertel's potential vorticity, Q, and winds on the 850 K isentropic surface for 5 December 1982. The units are 10^{-4} K m^2 kg^{-1} s^{-1}. In these units, light shading is for values between 1and 2, and heavy shading is for values between 3 and 5. Wind speeds can be read using the scale indicated at the bottom left of the figure. The frame surrounding the figure is tangent to latitude 20°N.

Figure 2. Polar-stereographic map of geopotential height (units: dam) at 100 mb on 5 December 1982. The frame surrounding the figure is tangent to latitude 20°N.

Figure 3. As for Fig. 1 but for 13 December 1982.

Figure 4. As for Fig. 1 but for 18 December 1982.

Figure 5. As for Fig. 1 but for 30 December 1982.

Figure 6. As for Fig. 2 but for 30 December 1982.

Figure 7. Graphs of areas A(Q) (units: 10^{12} m^2) as described in the main text. Each curve is labelled with the corresponding value of Q in units of 10^{-4} K m^2 kg^{-1} s^{-1}. The curves are for the northern hemisphere from July 1982 to June 1983. The equivalent latitude, marked on the right-hand axis, is calculated as the latitude circle that encloses the area marked on the left-hand axis.

Figure 8. Latitude-time section of zonal-mean, geostrophic wind (units: m s^{-1}) at 1 mb for December 1982 and the first half of January 1983.

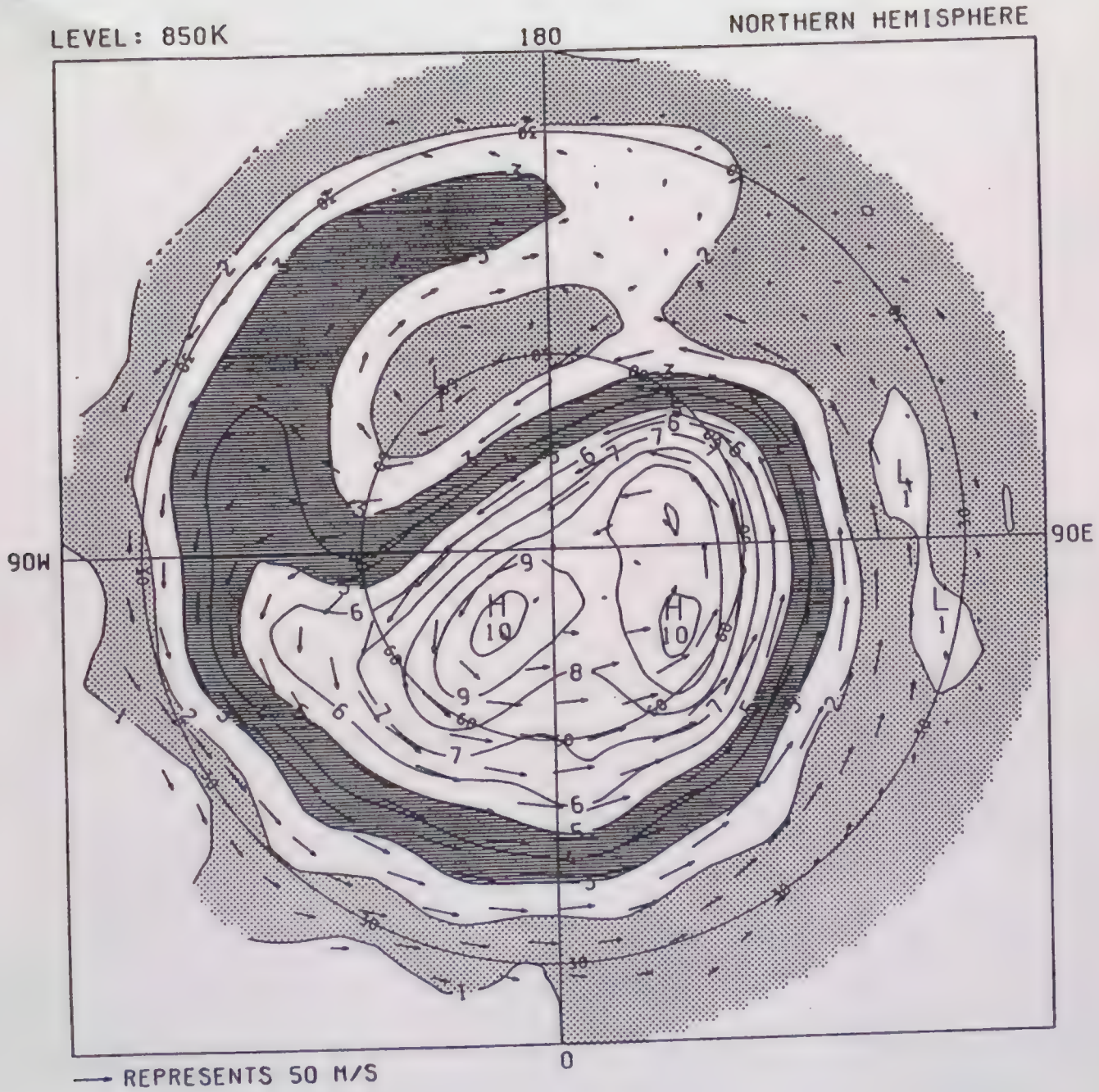

LEVEL: 850K
180
NORTHERN HEMISPHERE
90W
90E
0
H
10
L
1
REPRESENTS 50 M/S

Fig. 1

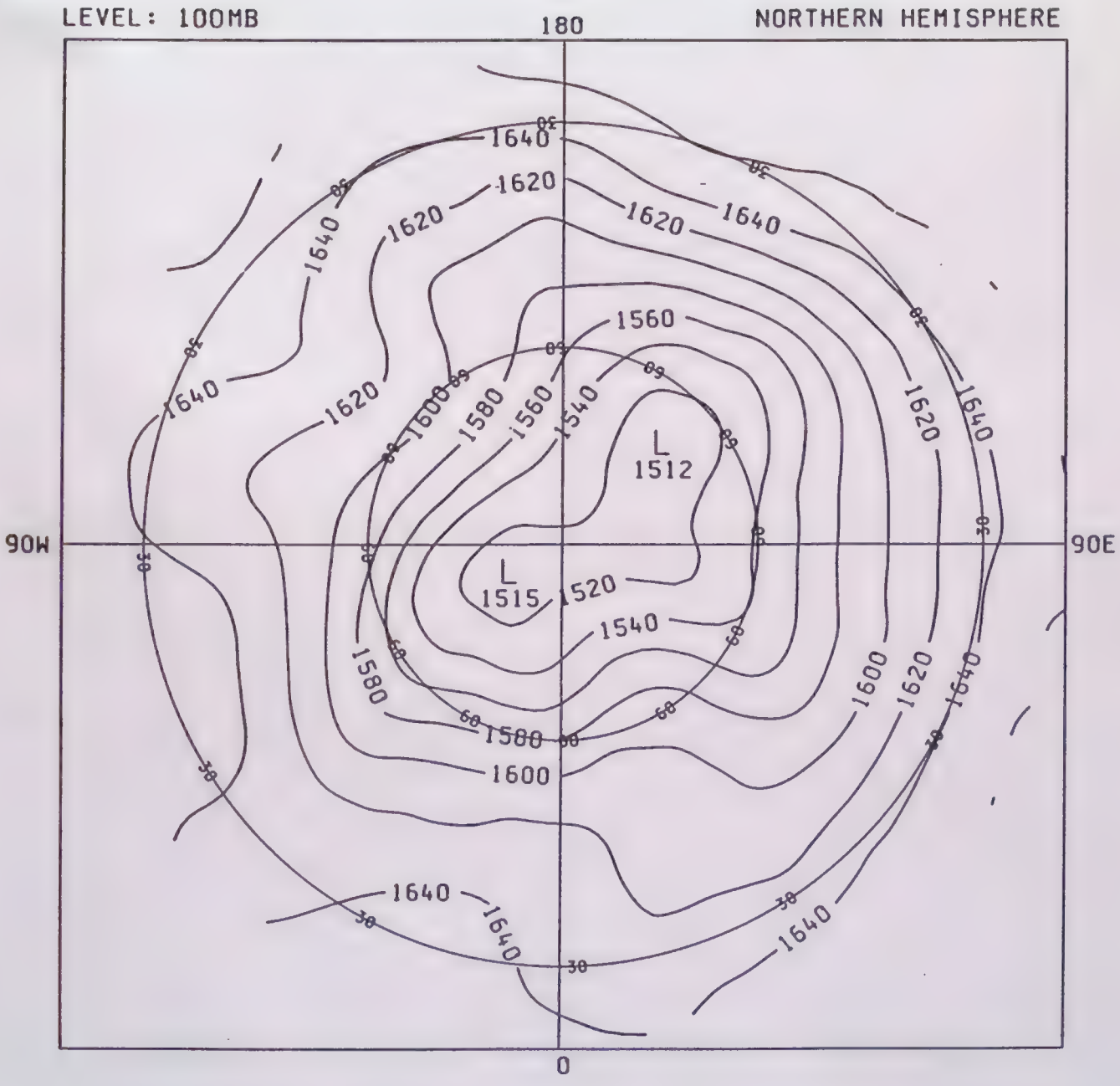

LEVEL: 100MB
180
NORTHERN HEMISPHERE
90W
90E
0
L
1512
L
1515
1520
1540
1560
1580
1600
1620
1640

Fig. 2

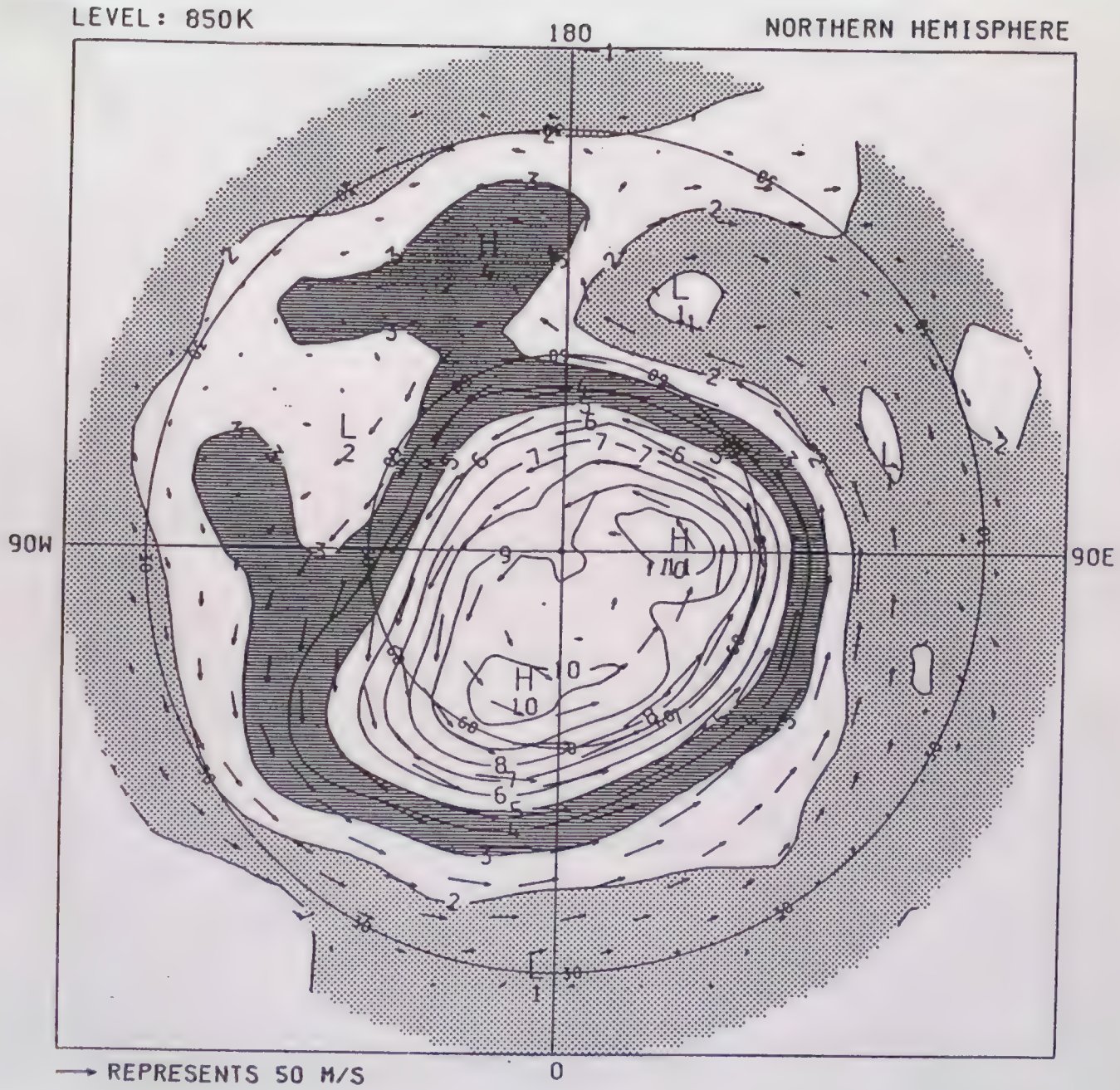

LEVEL: 850K
NORTHERN HEMISPHERE
180
90W
90E
0
H
L
REPRESENTS 50 M/S

Fig. 3

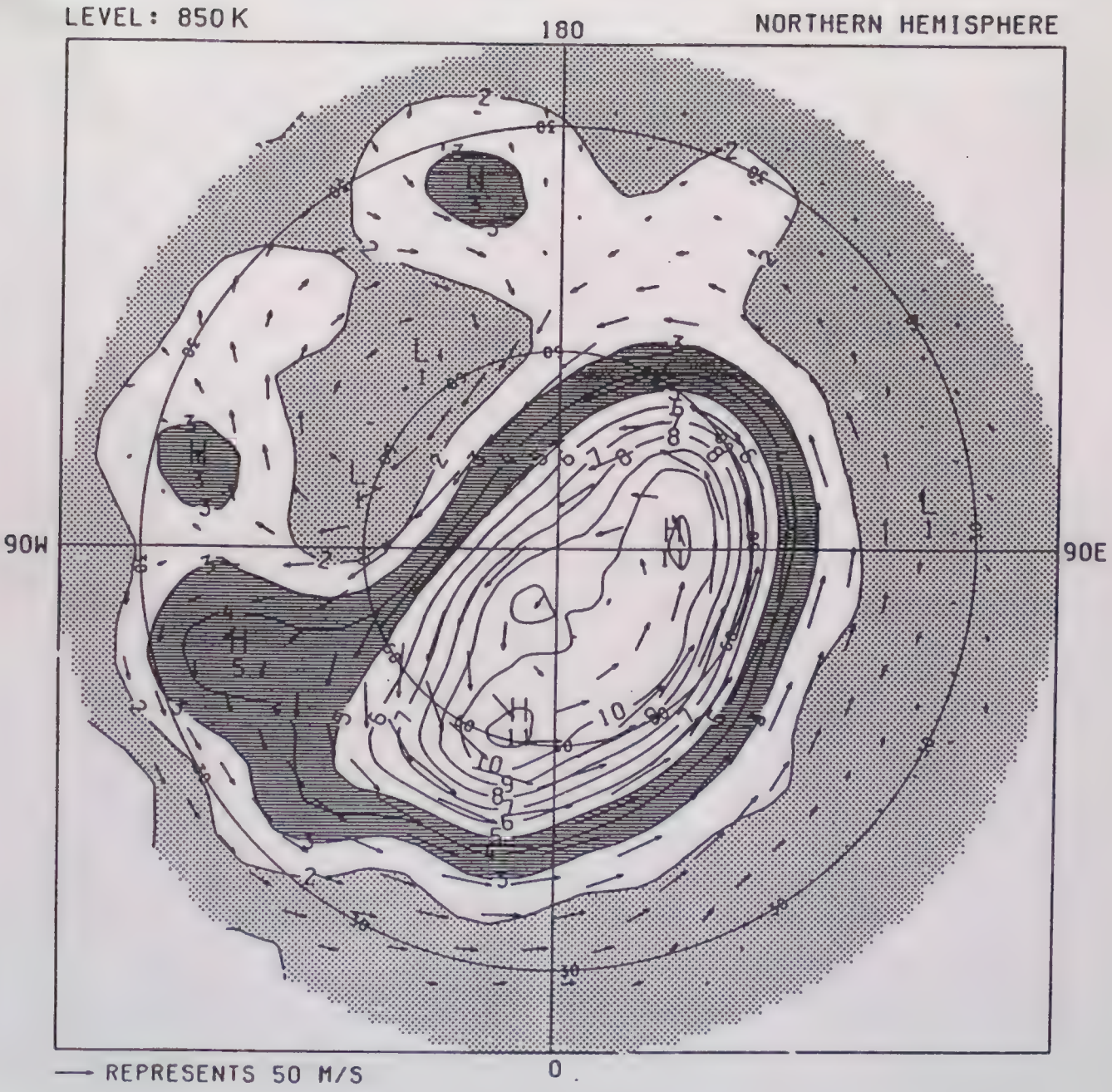

LEVEL: 850 K
NORTHERN HEMISPHERE
180
90W
90E
0
REPRESENTS 50 M/S

Fig. 4

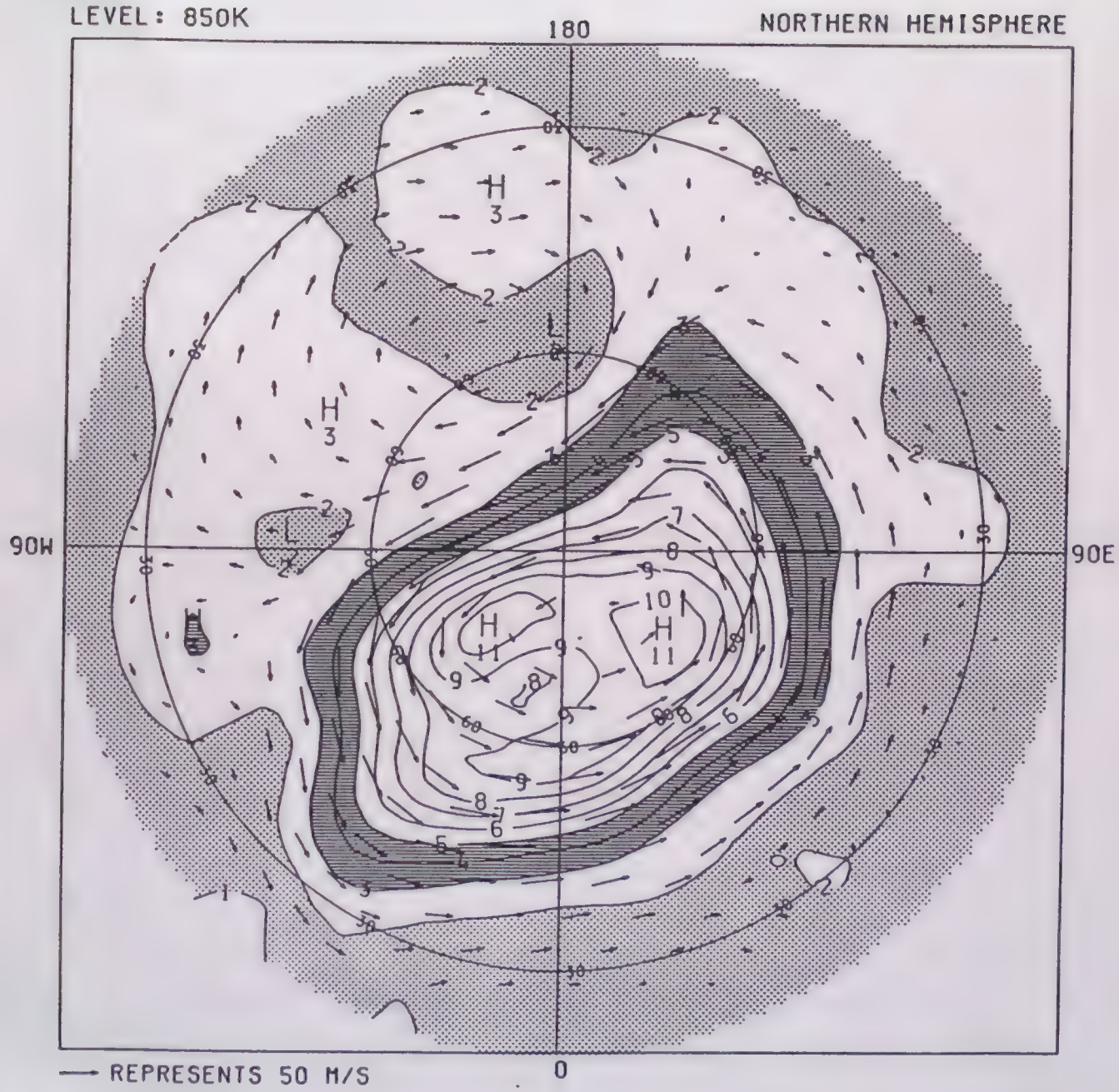

LEVEL: 850K
180
NORTHERN HEMISPHERE
90W
90E
0
REPRESENTS 50 M/S

Fig. 5

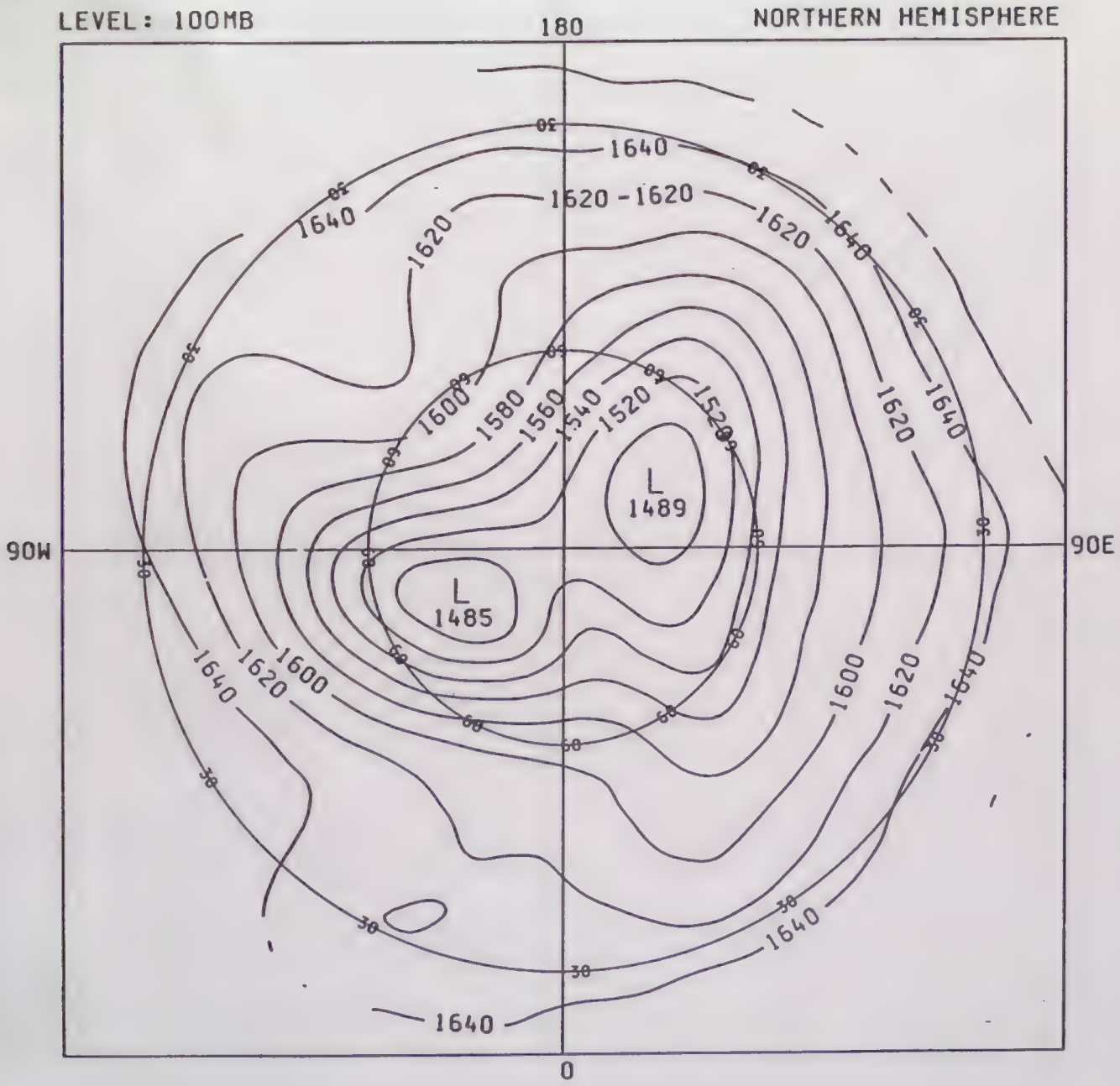

LEVEL: 100MB
180
NORTHERN HEMISPHERE
90W
90E
0
L
1489
L
1485
1640
1620
1600
1580
1560
1540
1520
30
60

Fig. 6

Fig. 7

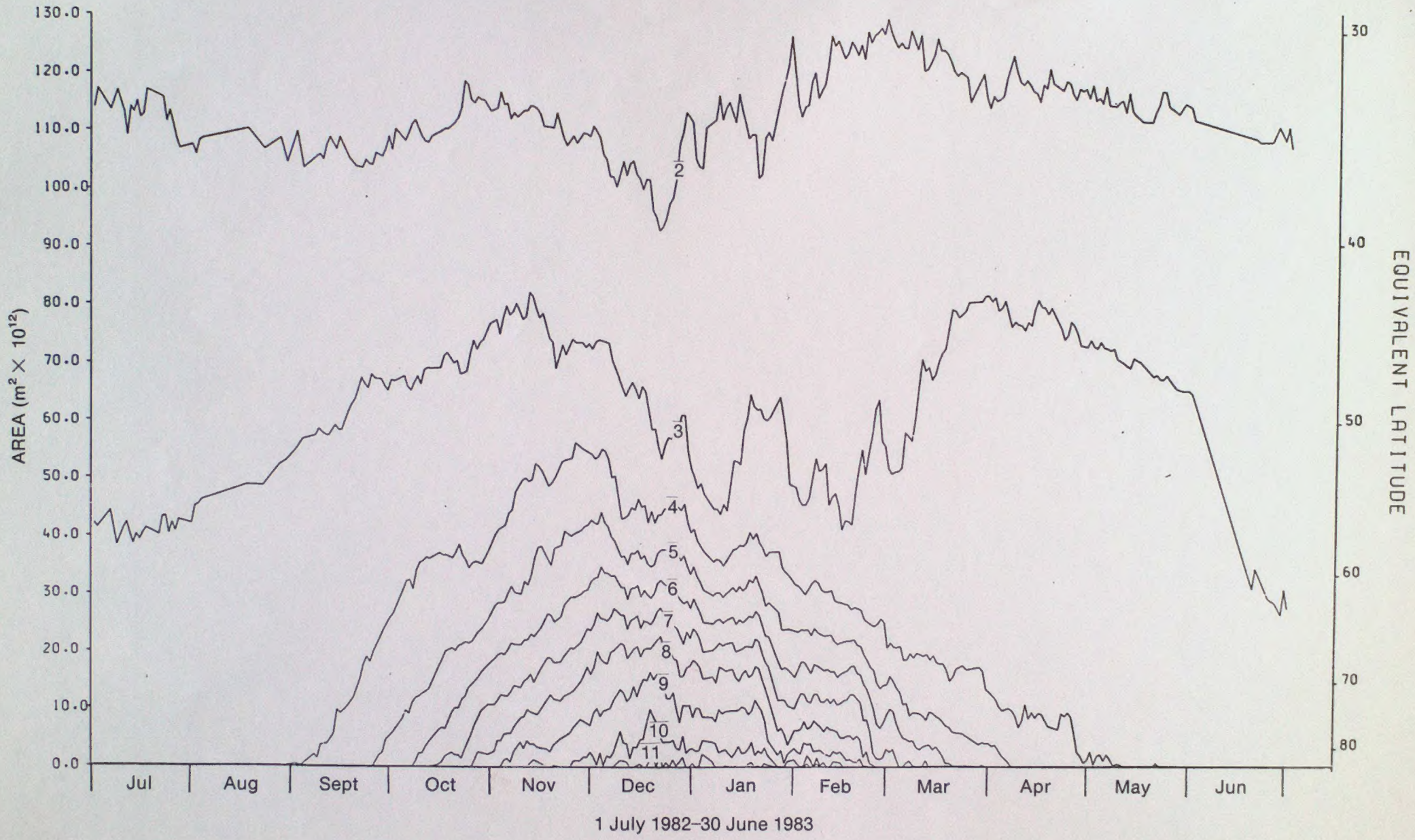
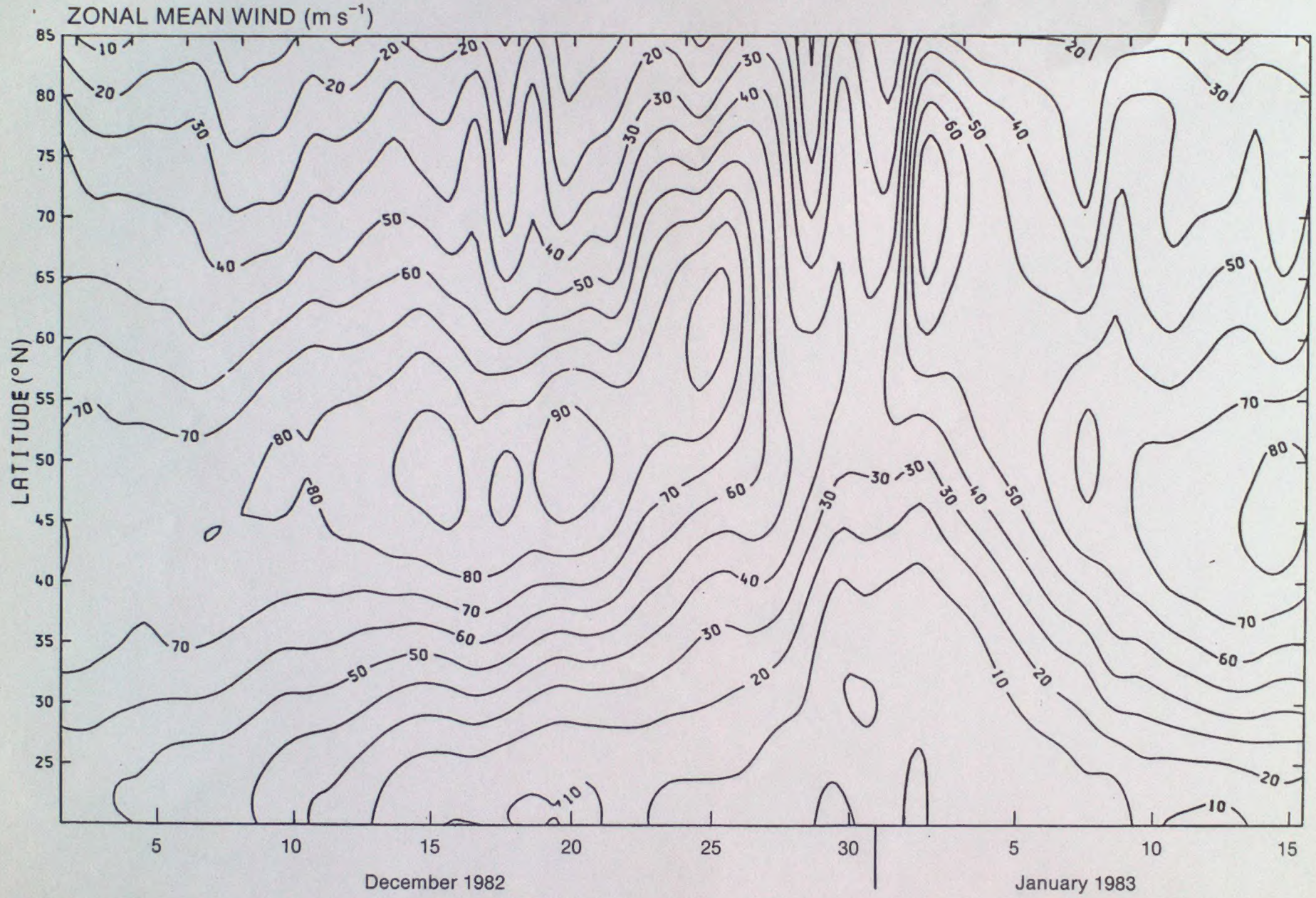

AREA ($m^2 \times 10^{12}$)
130.0
120.0
110.0
100.0
90.0
80.0
70.0
60.0
50.0
40.0
30.0
20.0
10.0
0.0
EQUIVALENT LATITUDE
30
40
50
60
70
80
2
3
4
5
6
7
8
9
10
11
Jul
Aug
Sept
Oct
Nov
Dec
Jan
Feb
Mar
Apr
May
Jun
1 July 1982–30 June 1983

Fig. 8


ZONAL MEAN WIND (m s⁻¹)
LATITUDE (°N)
85
80
75
70
65
60
55
50
45
40
35
30
25
5
10
15
20
25
30
5
10
15
December 1982
January 1983

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